The Colorado River extensional corridor in southwestern North America is one of Earth's most highly extended regions of continental crust. The central part of the belt includes three imbricate, regionally northeast-dipping extensional detachment faults. The Plomosa detachment fault in the northern Plomosa Mountains in western Arizona, the middle of the three faults, dips northeastward beneath the giant Harcuvar metamorphic core complex. Approximately 1 km of lower Miocene clastic sediments, lava flows, and rock-avalanche breccias were deposited in the northern Plomosa Mountains before initiation of the Plomosa detachment fault and division of the strata into two basins with different stratal accumulations following breakup. Both the detachment-fault lower plate and upper plate were then broken and tilted by normal faults. The upper plate was fragmented into numerous fault blocks and its extension-parallel width was approximately doubled. Application of critical-taper theory to delayed basin fragmentation suggests that southwestward tilting of the land surface and underlying normal faults led to normal-fault initiation and wedge breakup.

A seismic-reflection profile northeast of the northern Plomosa Mountains reveals strong, southwest-dipping reflectors that project up dip to metasedimentary tectonites in the southern Buckskin Mountains in the Harcuvar core complex. Restoration of displacement on the Buckskin and Plomosa detachment faults aligns the reflectors and tectonites with a Mesozoic shear zone in the footwall of the Plomosa detachment fault. In this restoration the combined shear zone dips northeastward rather than southwestward and projects up dip to the folds and thrusts exposed to the south and west of the northern Plomosa Mountains. This zone is interpreted as a segment of the Mesozoic Maria fold-and-thrust belt that influenced the geometry of younger detachment faults.

The southwest-tilted, mylonitic lower plate of the Plomosa detachment fault includes, at its northern end, Orocopia Schist, which is a Cretaceous subduction complex that is better known from locations farther southwest and closer to the continental margin. Restoration of tectonic extension suggests that Orocopia Schist extends under the Harcuvar core complex and that a buoyant crustal root inherited from Cretaceous thrusting could not have been the cause of core-complex uplift unless the schist was emplaced by a mechanism other than subduction underplating. We propose that the rolling-hinge detachment-fault model combined with a highly mobile deep crust could account for Harcuvar core-complex genesis without a buoyant crustal root.
The northern Plomosa Mountains, located in the central part of the Colorado River extensional corridor in western Arizona (Figs. 1, 2), consist of a structurally and lithologically complex assemblage of rock types separated into upper and lower plates by the regionally northeast-dipping Plomosa detachment fault (Fig. 3). This fault is the middle of three imbricate extensional detachment faults that are the dominant normal faults in this part of the extensional corridor (Spencer and Reynolds, 1991). New geologic mapping of the northern Plomosa Mountains (Spencer et al., 2015; Strickland et al., 2017b) combined with geochronologic data has clarified the early Miocene structural and stratigraphic evolution of the range and yields new insights into underlying processes of extensional tectonism across this part of the Colorado River extensional corridor.

This report describes the structure of the northern Plomosa Mountains and western Bouse Hills and presents a reconstruction of Miocene tilting and extension. It outlines the stratigraphy of tilted Miocene strata, describes the similarities and differences between fault blocks, and uses this information to infer details of extensional basin genesis and evolution. This outline of the chronology and kinematics of extension and basin genesis is followed by an evaluation of features and processes that influenced or controlled the structural and stratigraphic evolution of extension in this part of the Colorado River extensional corridor. Geodynamic issues addressed are follows: (1) Stable extension of the upper plate of the Plomosa detachment fault was followed by structural fragmentation after most sedimentation, a transition that is explained with critical-taper theory (Dahlen, 1984; Xiao et al., 1991). (2) Reconstruction of extension, along with correlation of structures and rock units in surrounding areas and interpretation of a seismic-reflection profile, leads to derivation of a reconstructed cross section that reveals previously unrecognized correlations of Mesozoic thrust-related structures and mylonitic metasedimentary tectonites. Identification of this dismembered thrust zone and its regional extent is consistent with the theory that weak carbonates buried by thrusting within dominantly quartzofeldspathic crust influenced the initial geometry of younger detachment faults, as proposed by Singleton et al. (2018). (3) The presence of latest Cretaceous subduction complex rocks below the Plomosa detachment fault (Strickland et al., 2017b) and at Cemetery Ridge 60 km to the southeast (Haxel et al., 2015) strongly suggests that Laramide (ca. 70–40 Ma) low-angle subduction removed the hypothetical crustal welt produced by older thrust faulting in the Maria fold-and-thrust belt and that crustal-root buoyancy could not have been the driving force for Harcuvar core-complex uplift as proposed by Coney and Harms (1984) and Spencer and Reynolds (1990b).
Figure 2. Geologic map of the Plomosa Mountains area showing location of map figures, seismic-reflection profile of Figure 9, and inferred locations of breakaways for major extensional detachment faults. “Pz” indicates metasedimentary tectonites at Battleship Peak derived from Paleozoic protoliths. Note that placement of the Plomosa Pass–Copper Peak breakaway ~10 km southwest of the trace of the Copper Peak detachment fault reflects the location of the fault trace farther west in the western Moon Mountains (Knapp, 1993).
Figure 3. Geologic map of the northern Plomosa Mountains and western Bouse Hills. Points A, B, and C (in red) are used as markers for calculation of total extension. Simplified from Spencer and Reynolds (1990a), Spencer et al. (2015), and Strickland et al. (2017b). See Figure 2 for location.
Geologic Setting

The Mojave-Sonoran Desert region of western Arizona and southeastern California, including the Colorado River extensional corridor, is within the transcontinental arch of cratonic North America, an area characterized by a thin platformal Paleozoic stratal section like that of the Grand Canyon (Stone et al., 1983; Sloss, 1988; Reynolds and Spencer, 1989). Crystalline basement with typical Arizona ages of ca. 1.4 Ga and ca. 1.7 Ga underlies the Paleozoic strata in southwestern North America (Richard et al., 2000). Lower Mesozoic strata were locally affected by deformation before eruption of widespread Jurassic felsic volcanic rocks, emplacement of granitoids, and local tectonic shortening (Reynolds et al., 1989; Tosdal et al., 1989; Boettcher et al., 2002; Tosdal and Wooden, 2015). Jurassic magmatism across southern Arizona and southeastern California was followed by latest Jurassic rifting and associated deposition of the lower part of the McCoy Mountains Formation (Tosdal and Stone, 1994; Spencer et al., 2011). After a relatively quiescent period, severe Late Cretaceous thrust faulting and related deformation produced the east-west–trending Maria fold-and-thrust belt in approximately the same area as the older McCoy rift basin and led to deposition of foredeep clastic deposits of the upper McCoy Mountains Formation (Hamilton, 1982; Reynolds et al., 1986; Laubach et al., 1989; Spencer and Reynolds, 1990b; Tosdal, 1990; Barth et al., 2004).

Late Cretaceous deformation and post-kinematic, ca. 80–70 Ma plutonism were followed by early Cenozoic erosion within a Cordilleran highland as indicated by the complete absence of lower Cenozoic sedimentary or volcanic rocks. During this time, low-angle subduction caused magmatism to shift inland to Colorado and New Mexico, while sediments shed northeastward from the Cordilleran highland accumulated on the adjacent area now uplifted to form the Colorado Plateau (Cather and Johnson, 1984; Young, 2001; Potochnik, 2001). Slab steepening and “melting back” associated with subduction of progressively younger oceanic lithosphere led to a resurgence of magmatism in the Mojave–Sonora Desert area at ca. 25 Ma and was associated with the beginning of extensional collapse of the Cordilleran orogen in western Arizona (Severinghaus and Atwater, 1990; Spencer et al., 1995; Sonder and Jones, 1999). The extreme extension in the central part of the Colorado River extensional corridor has been attributed to exceptional preexisting crustal thickness due to shortening within the Maria fold-and-thrust belt (Coney and Harms, 1984; Spencer and Reynolds, 1990b). This interpretation, however, is difficult to reconcile with latest Cretaceous to Paleocene subduction complex (Orocopia Schist) that replaced thickened crust beneath the central Colorado River extensional corridor before the onset of extension (Jacobson et al., 2011, 2017; Haxel et al., 2015; Strickland et al., 2017b).

**GOELOLOGY OF THE NORTHERN PLOMOSA MOUNTAINS AND ADJACENT AREAS**

The northernmost Plomosa Mountains consist of an ~15-km-long, southwest-tilted fault block overlain by the gently dipping Plomosa detachment fault and its extended upper plate (Scarborough and Meader, 1981, 1983; Spencer et al., 2015). Lower Miocene sedimentary and volcanic rocks resting depositionally on crystalline rocks of the footwall block dip 20°–70° to the southwest (south end of Fig. 3). Tilting of the footwall block presumably occurred during displacement on one or more structurally deeper, northeast-dipping normal faults that are concealed or not clearly identified within the Miocene volcanic field directly to the south (Fig. 2; Miller, 1970). The structurally lowest of these inferred faults, along with the Plomosa detachment fault and the structurally higher Buckskin-Rawhide detachment fault, make up the three major imbricate extensional faults at this section of the Colorado River extensional corridor (Spencer and Reynolds, 1991). The following description of structures and rocks is organized from southwest to northeast and is intended to document the structural geology and geologic history of these three major normal faults and associated features in the northern Plomosa Mountains and adjacent areas.

**Plomosa Pass–Copper Peak Breakaway**

The lower Miocene hornblende dacite at the south end of the map area of Figure 3 is part of a volcanic field that extends southward to Plomosa Pass (Fig. 2). The volcanic field is broken by northwest-striking normal faults and tilted to the southwest, but bedding dips are highly variable in some areas, probably because of the presence of eruptive centers and volcanic landforms (Miller, 1970; Sherrod et al., 1990). Farther south, the southern Plomosa Mountains consist of Proterozoic, Paleozoic, and Mesozoic rock units that are cut by Mesozoic thrust faults and younger high-angle faults (Miller, 1970; Miller and McKee, 1971; Sherrod et al., 1990; Richard et al., 1993; Richard and Spencer, 1994). These rocks are overlain by scattered masses of Miocene volcanic and sedimentary rocks that are slightly to moderately tilted, but they and underlying pre-Cenozoic rocks are not generally broken into tilted fault blocks and are not significantly extended. The contrast in extensional faulting north and south of Plomosa Pass has been inferred to reflect a concealed, northeast-dipping normal fault, with characteristic southwest tilting of volcanic rocks in its hanging-wall block (Spencer and Reynolds, 1991). It also marks the southwestern boundary of the Colorado River extensional corridor.

The north-south–trending Dome Rock Mountains and their northern continuation as the eastern Moon Mountains (Figs. 1, 2) form a 40-km-long fault block bounded on their eastern and western sides at least partially by outward-dipping normal faults (Hauser et al., 1987; Spencer et al., 2016a) and at the northern and southern ends by outward dipping low-angle normal faults (Tosdal and Stone, 1994; Knapp, 1993; Fig. 4). The Dome Rock Mountains are analogous to the southern Plomosa Mountains in their lack of significant internal extension (Tosdal and Stone, 1994; Johnson et al., 2017) and are similarly placed southwest of the edge of the Colorado River extensional corridor (Figs. 1, 2). The Copper Peak detachment fault at the northern end of the eastern Moon Mountains dips gently northeastward beneath the adjacent Copperstone open-pit mine (Fig. 4). The Copperstone deposit is the largest of a group of Fe-Cu-Au deposits produced by oxidizing hydrothermal basin brines that flowed along active detachment faults in the Colorado River extensional corridor (Wilkins and Heidrick, 1982; Spencer and Welsey, 1986, 1989; Duncan, 1990a; Salem, 1993). Northeast-dipping normal faults at Copper Peak and inferred in the Plomosa Pass area, both of which mark the southwest margin of the Colorado River extensional corridor, are here referred to as the “Plomosa Pass–Copper Peak breakaway” (Fig. 2).

**Northern Plomosa Mountains**

**Crystalline Rocks**

The footwall block of the Plomosa detachment fault consists of igneous and metamorphic rocks with depositionally overlying, lower Miocene strata at the southern end of the fault block. Granitic and gneissic rocks directly below the tilted Miocene strata are faulted over folded and sheared Paleozoic and Mesozoic metasedimentary rocks of the generally south-dipping Quinn Pass shear zone (Fig. 3). West of Quinn Pass, the metasedimentary rocks form an east-west–trending belt faulted between adjacent foliated granitic and gneissic rocks. Folation in metasedimentary rocks is subparallel to foliation in adjacent crystalline rocks and to the fault contacts that separate all three assemblages. Tight to isoclinal folds within
the Quinn Pass shear zone, typically involving Permian Kaibab Limestone and Triassic clastic units, have axes parallel to the margins of the shear zone (Stoneman, 1985a, 1985b; S.J. Reynolds, unpublished data). The belt is broken by a north-trending high-angle fault at Quinn Pass and is much thicker to the east. Eastern exposures are highly deformed, with four sets of superimposed folds, the youngest of which trend southeastward but are otherwise similar in structural style and geometry to the folds west of Quinn Pass (Steinke, 1996, 1997). Crystalline rocks directly north of the Quinn Pass shear zone consist of thinly layered, fine-grained quartzo-feldspathic gneiss and mafic gneiss and a small intruding diorite body.

Micaceous quartzo-feldspathic schist correlated with the Orocopia Schist forms the northern part of the exposed footwall (Fig. 3; Strickland, 2017, 2017b). This schist contains sparse lenses of actinolite and rare talc schist. Plagioclase poikiloblasts commonly contain graphite inclusions. Local hornblende amphibolite layers interpreted as metabasalt include small bodies of recrystallized fine-grained quartz with ferromanganiferous garnet that are interpreted as metachert. Detrital-zircon geochronologic analyses of schist derived from sandstone indicate a maximum depositional age of 70–75 Ma (Strickland, 2017; Seymour et al., 2018). All these features support correlation with Orocopia Schist, which is interpreted as subduction complex underplated beneath southwestern North American during Laramide low-angle subduction (Grove et al., 2003; Jacobson et al., 2011, 2017; Haxel et al., 2015; Chapman, 2016).

Orocopia Schist and adjacent gneissic rocks are intruded by a 22–23 Ma, compositionally expanded suite of granitoids including layered tabular bodies of leucocratic biotite tonalite and granodiorite and hornblende-biotite diorite (Strickland, 2017). The intrusive complex, Orocopia Schist, and nearby crystalline rocks are overprinted by a gently dipping mylonitic fabric with a strong lineation that is better developed in more northern exposures. Average plunge and trend of mylonitic lineations is 9° toward 220°, with numerous top-northeast shear-sense indicators (Strickland et al., 2017a, 2017b, 2018).

Crystalline rocks in tilted fault blocks above the Plomosa detachment fault consist of coarse, porphyritic biotite granite and, in one fault block, foliated Mesozoic leucogranite that intrudes numerous faulted masses of Paleozoic carbonate. In three fault blocks, granitic rocks are thrust over deformed Paleozoic and Mesozoic metasedimentary rocks (Fig. 3). The style of deformation in these rocks is generally similar to that of the Maria fold-and-thrust belt elsewhere (e.g., Hamilton, 1982; Reynolds et al., 1986; Boettcher et al., 2002).
Miocene Sedimentary and Volcanic Rocks

Both the lower plate of the Plomosa detachment fault and the multiple tilted fault blocks of the upper plate include ~1-km-thick sections of southwest-tilted lower Miocene strata (Figs. 3, 5, 6). The northwestern Bouse Hills form a tilted fault block with similar stratigraphy and are included here as the structurally highest and northeasternmost tilted fault block above the Plomosa detachment fault. Two composite units are present in all sections (numbered 1 and 2 in Fig. 6). A lower composite unit consists of basal conglomerate and arkosic sandstone that grades upward into sandstone, siltstone, and lacustrine limestone. This is overlain by an upper composite unit of mafic lava flows and rock-avalanche breccia sheets and lenses. The breccias were derived largely from Paleozoic carbonates and quartzites and, less commonly, granitoids of inferred Jurassic or Proterozoic age. While these two composite units are similar across all fault blocks, overlying strata (numbered 3 in Fig. 6) are dissimilar in the footwall and hanging-wall blocks of the Plomosa detachment fault.

In the footwall block, composite unit 2 is overlain by a thick section of hornblende dacite. In three hanging-wall tiltblocks in the central area of Figure 3, composite unit 2 is overlain by conglomerate and sandstone, with limestone in one fault block. At the top of two of these three stratal sections, conglomerates contain sparse mylonitic and chloritic breccia clasts presumably derived from the lower plate of the detachment fault. Two northern sections do not preserve strata above composite unit 2. Finally, section A (Fig. 6) in the Bouse Hills is part of the footwall of the Buckskin detachment fault, and the lower part of the section may have displaced correlatives in the Artillery Mountains and Date Creek basin ~60 km to the northeast (Spencer et al., 2016b).

Thermochronology of Extension

Seven samples of crystalline rocks from a north-south transect in the lower plate of the Plomosa detachment fault yielded zircon and apatite fission-track dates that reveal a laterally varying early Miocene cooling
history (Foster and Spencer, 1992). Six dates from four samples of the southern part of the footwall block are 17–23 Ma, while five dates from three samples of northern footwall rocks are 15–19 Ma (Fig. 7). Cooling dates generally decrease northward. The Mudersbach granite near the southern samples, dated at 20.4 ± 2.0 Ma by U-Pb (Steinke, 1996), was intruded at about the same time as cooling of fission-track samples through annealing temperatures, as was a small diorite body dated by U-Pb zircon at 20.46 ± 0.15 Ma (Strickland et al., 2017b; Fig. 3). In contrast, the 22–23 Ma intrusive complex in the northern Plomosa Mountains is older than the 15–19 Ma cooling dates.

Bouse Hills

As shown in Figure 2, the Bouse Hills consist of a western area of lower Miocene volcanic and minor sedimentary rocks, a central area of Proterozoic granitic and gneissic rocks, and an eastern granitic pluton dated at 21.7 ± 0.2 Ma (U-Pb zircon date from Singleton et al., 2014). In the northwestern Bouse Hills, pre-Miocene granitic rocks are depositionally overlain by basal arkosic sandstone and limestone, which in turn are overlain by mafic lava flows and rock-avalanche breccias (Figs. 3, 5, 6; Spencer and Reynolds, 1990a). The entire section is tilted steeply to the southwest. Biotite from a tuff near the base of the section yielded an 40Ar/39Ar date of 22.87 ± 0.10 Ma (average of three step-heating plateau dates from Singleton et al., 2014). Tilted strata are overlain by generally subhorizontal felsic to intermediate volcanic rocks, which in turn are overlain by mafic lava flows and basaltic lava flows. Three K-Ar dates from the upper tuff (biotite and hornblende) and basalt (whole rock) have a weighted mean of 19.7 ± 1.0 (2σ) Ma (Fig. 8; Spencer et al., 1995; weighted mean and uncertainty calculated by method of Long and Rippeteau, 1974). A transect of five (U-Th)/He apatite dates from crystalline rocks from below the basal arkose to the eastern end of the Bouse Hills indicates low-temperature cooling from west to east at ca. 21–17 Ma (Singleton et al., 2014).

Cactus Plain and the Southern Buckskin Mountains

The Battleship Peak area in the southern Buckskin Mountains is underlain on its southwestern flank by marble, calc-silicate, quartzite, and felsic intrusions that are strongly foliated and mylonitic (Fig. 2; Marshak and Vander Meulen, 1989; Singleton et al., 2018). These tectonites, largely derived from Paleozoic sedimentary units, are sheared against underlying crystalline rocks, and a basal Cambrian quartzite and depositional contact are absent. A seismic-reflection profile across Cactus Plain northeast of the northern Plomosa Mountains, produced as part of a much longer...
COCORP seismic profile across most of Arizona (Hauser et al., 1987), revealed a 2–3-km-thick set of southwest-dipping, strong seismic reflectors that project up dip into layered tectonites and gneisses in the southern Buckskin Mountains (Fig. 9). These strong reflectors are underlain by weaker reflections that also dip southwest. All these reflectors, except for those deeper than ~3 seconds two-way travel time, appear to be cut by a thin set of northeast-dipping reflectors (22° apparent dip at 5 km/s, at the strike of the profile) that project up dip to the southwest toward the trace of the Plomosa detachment fault.

CHRONOLOGY AND KINEMATICS OF SEDIMENTATION AND FAULTING

Chronology of Magmatism and Extension

The 22.87 ± 0.10 Ma tuff bed near the base of tilted strata in the western Bouse Hills indicates that basin genesis in the northern Plomosa Mountains area began at ca. 23 Ma (Singleton et al., 2014; Fig. 6). A U-Pb date of 21.07 ± 0.15 Ma was determined for a tuff bed just above the base of tilted strata resting on the footwall block of the Plomosa detachment fault (Strickland et al., 2018; Fig. 6). These dates suggest that sediments onlapped southwestward across the landscape at ca. 23–21 Ma. Intraformational activity in the region began during or shortly after initial basin subsidence, as represented by the following: (1) the 20.4 ± 2.0 Ma Mudersbach pluton east of Quinn Pass (Fig. 3; Steinke, 1996); (2) the 20.46 ± 0.15 Ma diorite northwest of Quinn Pass (Strickland et al., 2017b); (3) the 22–23 Ma intrusive complex in the northern Plomosa Mountains (Strickland et al., 2017b); (4) the 21.7 ± 0.2 Ma pluton that forms the eastern Bouse Hills (Singleton et al., 2014); and (5) the Swansea plutonic suite in the lower plate of the Buckskin-Rawhide detachment fault that yielded four U-Pb dates of 21–22 Ma (Bryant, 1995; Bryant and Wooden, 2008; Singleton and Mosher, 2012).

Low-temperature thermochromometers suggest that cooling related to tectonic exhumation continued for several million years after the ca. 20–23 Ma magmatic pulse. Zircon fission tracks are annealed on geologic timescales at ~200 °C (Naeser, 1981; Green et al., 1989; Brandon et al., 1998), while apatite fission tracks are annealed above ~110 °C and anneal at progressively slower rates down to ~60 °C (Hurford, 1986; Gleadow and Fitzgerald, 1987). Five fission-track dates from three samples of the mylonitic, northernmost part of the Plomosa detachment footwall indicate that cooling through annealing temperatures of zircon and apatite occurred at ca. 15–19 Ma (Fig. 7). Similarly, emplacement of the 21.7 ± 0.2 Ma pluton in the eastern Bouse Hills was followed by low-temperature cooling in crystalline rocks of the Bouse Hills and nearby southern Buckskin Mountains at ca. 16–19 Ma as indicated by 8 (U-Th)/He dates on apatite (Singleton et al., 2014, average dates from 3 to 4 dated aliquots for each sample). In both areas, this chronology is interpreted to indicate the ca. 15–19 Ma timing of cooling due to tectonic exhumation to within 2–4 km of Earth’s surface. In contrast, the six fission track dates from the southern part of the lower plate of the Plomosa detachment fault, which range from ca. 17 to 23 Ma, are not clearly distinguishable from cooling following heating during emplacement of the 20.4 ± 2.0 Ma Mudersbach granite and magmatism associated with emplacement of a small, 20.46 ± 0.15 Ma diorite (Strickland et al., 2017b). The apatite and zircon fission-track dates from each location are concordant within error, indicating that the footwall rocks experienced geologically rapid cooling from above ~250 °C to below ~60 °C. We conclude that cooling related to tectonic exhumation by two imbricate detachment faults was under way in both the northernmost Plomosa Mountains and Bouse Hills–southern Buckskin Mountains at ca. 16–19 Ma, although exhumation most likely began earlier.

Miocene Basin Evolution

As indicated above, most Miocene strata in the northern Plomosa Mountains and Bouse Hills can be divided into two composite units (Fig. 6). The lower composite unit, consisting of generally fining-upward conglomerate, sandstone, and lacustrine limestone, represents basin genesis in a region of Cretaceous and older crystalline rocks within an eroding Paleogene highland. The geometry of the basin is not obvious, but the rounded character of conglomerate clasts in some basin conglomerates suggests that the developing basin captured streams of moderate size. Clastic strata are interbedded with limestone, possibly indicating laterally migrating depocenters or transient closed-basin conditions as might be expected for an active extensional basin.

The upper composite unit (unit 2 in Fig. 6) consists primarily of mafic lava flows but includes intermediate to felsic lavas and tuffs and abundant interbedded rock-avalanche breccia sheets and lenses of Paleozoic and Mesozoic strata and granitoids (Spencer et al., 2015). In the simplest interpretation, mafic lava flows accumulated in the previously formed sedimentary basin, while substantial relief at a faulted margin of the basin yielded rock-avalanche breccias that fell from locations that were high enough for falling rock masses to shatter and acquire the velocity necessary for acoustic fluidization and sheet-like dispersal over gentle slopes (Melosh, 1979). This faulted basin margin could have been the Plomosa Pass–Copper Peak breakaway, with metamorphic and granitic rocks derived from the breakaway fault scarp. Composite units 1 and 2, however, do not thicken southwestward toward this inferred scarp, indicating that scarp genesis was not linked to simple half-graben subsidence and southwestward tilting.

The similarity of composite unit 2 in the footwall and hanging wall of the Plomosa detachment fault indicates that this detachment fault had not yet become active or did not produce enough displacement to influence the distribution of mafic lavas and rock-avalanche breccias. Overlying strata (composite unit 3 in Fig. 6) consist of different rock types in different areas and are most readily interpreted to reflect basin disruption by normal faulting. Specifically, the voluminous hornblende-dacite lava and related units that were deposited south of Quinn Pass on the footwall block of the Plomosa detachment fault are almost entirely absent from
fault blocks in the hanging wall (Fig. 6). Similarly, clastic sediments above composite unit 2 are completely absent south of Quinn Pass. This contrast is interpreted to indicate that significant displacement on the Plomosa detachment fault divided the basin represented by composite units 1 and 2.

The northern Plomosa Mountains include two areas of distinctive, poorly sorted conglomerate that contain sparse clasts of mylonitic crystalline rocks and chloritic breccia that are similar to crystalline rocks of the detachment-fault footwall (Fig. 6, columns C, D). The presence of mylonitic and chloritic breccia debris is interpreted to indicate that the mylonitic detachment-fault footwall was exposed at the time this conglomerate was deposited. Erosion and dispersal of clasts derived from the footwall of the Plomosa detachment fault must have occurred after footwall cooling through apatite fission-track annealing temperatures, indicating that this conglomerate is younger than ca. 16 Ma.

It is not apparent from measured bedding dips that stratigraphically higher strata are tilted less than lower strata, although pervasive disruption of all upper plate rock units obscures such possible relationships (Spencer et al., 2015). Regardless, most or all tilting occurred after all Miocene strata were deposited which would make tilting younger than ca. 16 Ma according to the chronology outlined above. This interpretation, and the ca. 19 Ma K-Ar dates of untilted volcanic rocks in the central Bouse Hills (Fig. 8), indicate that the untilted volcanic rocks in the Bouse Hills were too far east of the trace of the Plomosa detachment fault to be affected by wedge breakup and tilting, or that tilting in the two areas was diachronous, occurring before ca. 19 Ma in the Bouse Hills and after ca. 16 Ma in the northern Plomosa Mountains.

Plomosa Detachment-Fault Geometry

Lower Miocene strata resting on the footwall block of the Plomosa detachment fault south of Quinn Pass dip ~40°–70° southwestward, indicating that tilting of the fault and footwall block was in this range. Restoration of 60° of rigid fault-block tilting places northernmost exposures of the footwall at ~13 km depth below the basal Miocene nonconformity and at ~14 km below the top of the basalt and breccia unit. It is likely, however, that the Plomosa fault had a broadly listric form, flattening downward and not reaching such depths at its northern footwall exposures. This is suggested by the geometry of other core complexes in the area with similarly steep dips at the structural top of core complexes but including evidence that the structurally deepest parts of the core complexes were at shallower depths than indicated by simple block tilting (e.g., Spencer and Reynolds, 1991; Gans and Gentry, 2016). North of Quinn Pass the trace of the Plomosa detachment fault bends ~90°, most likely marking the northeast-trending axis of a synformal groove in the detachment fault but also possibly reflecting listric curvature of the detachment fault. If this curvature marks a 30° decrease in original fault dip, then the paleodepth of exhumed rocks at the north end of the footwall would be 9 km below the nonconformity and 10 km below the top of the basalt and breccia unit. We conclude that the paleodepth of the northern end of the exposed footwall was ~9–14 km below Earth’s surface before tectonic exhumation.

The ~11-km-long lateral ramp in the Plomosa detachment fault, where the fault dips eastward at ~10°–30°, is either an original lateral ramp along the flank of a corrugation, or the result of post-detachment eastward tilting in the footwall block of a concealed, hypothetical west-dipping normal fault west of the range. Note, however, the map pattern of tilted Miocene strata around the northern half of cross section D–D′ (Fig. 3). The upper three units thicken eastward or southeastward, forming a fanning pattern in map view. If the lateral ramp were the result of post-detachment, eastward tilting, the upper plate half-graben sequence would have been tilted eastward, about an axis at a high angle to the axis of the half graben, so that it would begin to resemble a half-graben cross section in map view. In this case, half-graben strata would be expected to thicken toward the detachment fault, which would have been beneath the deepest part of the half-graben fill. This is not the case, which suggests that the lateral ramp is an original feature and not due to post-detachment tilting. North of the lateral ramp the trace of the detachment fault bends westward and the fault dips northward. This bend appears to mark the top of a northeastward-trending antiformal corrugation.
Structural Reconstruction

The basal depositional contact of lower Miocene strata in the northern Plomosa Mountains and Bouse Hills provides a datum for determining the width of the pre-extension landscape and the total amount of extension. The composite cross section of Figure 10, derived from the detailed cross sections of Figure 5 projected onto a single profile, shows 6 upper-plate normal faults separating 7 tiltblocks. These faults are all represented as having a modern dip of 30°, although fault exposures are so poor that only one outcrop revealed fault dip (67° on a second-generation normal fault; Fig. 3). Fault-bedding intersection angles approaching 90° in cross sections probably do not represent initially near-vertical faults. These intersection angles likely resulted from tilting associated with displacement on the Plomosa detachment fault before initiation of ~45°–60° dipping, upper-plate normal faults that cut previously tilted strata. In addition, tilting could have been caused by more than one generation of normal faults. East of the central part of cross section B–B′ (Figs. 3, 10), a 90° curve in the trace of a normal fault is interpreted as the side of a scoop-shaped, gently dipping fault surface that was truncated by the younger normal fault with the 67° dip measurement.

Detailed mapping indicates that bedding strike and dip commonly vary by 10°–30° over ~50–500 m without obvious intervening structures (Spencer et al., 2015). This is attributed to general brittle structural disruption, especially close to the Plomosa detachment fault. This disruption indicates that fault blocks were not rigid and that complex to penetrative brittle deformation within fault blocks likely accommodated displacement at corners where upper plate normal faults intersect the detachment fault at gentle angles. We emphasize that the assumption of 30° fault dips shown in Figure 10 is an approximation that conceals multiple complexities that are poorly understood.

The width of the landscape that was buried by lower Miocene strata and later broken and distended by faulting can be approximated from the composite cross section of Figure 10. According to this simplified cross-section analysis, a landscape that was originally 9.6 km wide is now 20.7 km wide (separation of points “A” and “C” on Figs. 3, 5, and 10) and reflects ~116% extension. Calculation of extension between points “A” and “B” does not include the partially concealed half-graben between the Bouse Hills and northern Plomosa Mountains but yields essentially the same percentage extension (113%).

Normal-Fault and Basin Evolution

The earliest known extension in the Colorado River extensional corridor is represented by clastic strata at the base of a half-graben in the Artillery Mountains (Fig. 1) that contain a 26.28 ± 0.01 Ma tuff bed (Lucchitta and Suneson, 1993). The axis of the half-graben extends southeastward beneath Date Creek basin (Fig. 1) where the basin axis is the inferred location of the trailing edge of pre-Cenozoic basement of the upper plate of the Buckskin detachment fault (Spencer et al., 2016b). Figure 11 shows a schematic cross-section evolution through the major detachment faults in this section of the Colorado River extensional corridor. The lower Artillery Formation is shown adjacent to the Buckskin detachment breakaway in the Bouse Hills, with younger Plomosa composite units 1 and 2 extending northeastward into what is now Date Creek basin (Fig. 11A). In this cross section the Artillery Formation is projected from the Artillery Mountains southeastward along strike by 10 km (Singleton, 2015) to 20 km (Spencer et al., 2016b) to where it would be tectonically restored to a position adjacent to the Bouse Hills. Correlation of the middle Artillery Formation with composite unit 2 in the Bouse Hills is suggested by lithologic similarity, with both containing abundant mafic lava flows and rock-avalanche brecias of similar composition (Yarnold, 1994; Spencer et al., 2013).

Northern Plomosa basin genesis and creation of relief sufficient to yield rock-avalanche breccias was followed by movement on the Plomosa...
Above normal faults is possible (Fig. 12; see also Spencer et al., 2016b). Like dry sand, non-cohesive materials can support a surface slope up to an angle of repose. The only rock properties relevant to critical-taper theory in non-cohesive material over distances of kilometers to tens of kilometers. Forming a wedge above a normal or thrust fault can be approximated as a schematic representation.

**GEODYNAMICS OF EXTENSION**

**Wedge Breakup**

Critical-taper theory may offer insight into the causes of breakup of normal-fault blocks (Dahlen, 1984; Xiao et al., 1991). Critical-taper theory is based on the assumption that rocks in the brittle deformation regime forming a wedge above a normal or thrust fault can be approximated as a non-cohesive material over distances of kilometers to tens of kilometers. Like dry sand, non-cohesive materials can support a surface slope up to an angle of repose. The only rock properties relevant to critical-taper theory in extension (other than fluid pressure, which is ignored here) are the maximum surface slope that can be sustained and the minimum fault dip for normal slip (other than fluid pressure, which is ignored here) are the maximum angle of repose. The only rock properties relevant to critical-taper theory in non-cohesive material over distances of kilometers to tens of kilometers.

**Reactivation of a Thrust Zone**

A cross section through the Plomosa Mountains area (Fig. 13A) includes the simplified composite cross section of Figure 10 and extends southwestward and northeastward to adjacent areas (Fig. 2). The Oroopia Schist is projected laterally onto the cross section from exposures in the northern Plomosa Mountains (Strickland et al., 2017b). Also shown are the strong seismic reflectors beneath Cactus Plain (Hauser et al., 1987) which project up dip toward the Battleship Peak area and its layered metasedimentary tectonites (Marshak and Vander Meulen, 1989; Singleton

**Figure 11. Schematic cross-section evolution of tectonic extension in the northern Plomosa Mountains and adjacent areas.** The three major normal faults are named. Dotted lines represent faults at time of initiation. Points A and C (in red) represent basal Miocene contacts identified in Figures 3, 5, and 10. A second generation normal fault that crosses the central part of cross section line B–B′ on Figure 3 is shown cutting the first generation above the Plomosa detachment fault (Fig. 11C). Fault dips at depth are uncertain and could have been listric. Internal deformation of fault blocks to accommodate displacements at fault intersections is not shown in this schematic representation.

**Figure 12. Critical-taper representation of extensional evolution of the upper plate of the Plomosa detachment fault.** The parameters defining the stable-sliding region, $\phi = 13^\circ$ (maximum supportable surface slope) and $\phi_b = 3^\circ$ (minimum fault dip), represent the extensional Coulomb-wedge calibration of Spencer (2011). A wedge of rock exhibiting Coulomb behavior will slide stably down a fault ramp if surface slope and fault dip are within the stable-sliding region. Vector “a” represents tilting of a wedge with an initially horizontal upper surface above a 60°-dipping normal fault. Vector “b” is for a similar wedge initially above a 45° dipping normal fault. If the fault and wedge surface are tilted toward the fault breakaway in idealized half-graben behavior, the wedge configuration will reach the upper boundary of the stable-sliding field. Further tilting will cause internal wedge extension so that the representation on the plot does not exceed the boundary of the stable-sliding field (wedge configuration along curved red line leading to point “c”). The dotted lines represent the possible evolution if erosion and sedimentation are effective at impeding development of significant surface slope during tilting. These evolutionary pathways are inferred to approximate the evolution of the upper plate of the Plomosa detachment fault, with composite units 1 and 2 deposited during critical taper configurations represented by vectors “a” and “b.” The evolutionary pathways shown also may represent the pathway for initiation of the Plomosa detachment fault and perhaps other normal faults.
et al., 2018). The imbricate Plomosa detachment fault and the Plomosa Pass–Copper Peak breakaway fault are shown as separated by an additional normal fault that is a schematic representation of a more complicated situation. Some of the normal faults mapped by Miller (1970) in the volcanic field north of Plomosa Pass are southwest-side-down, as shown on Figure 2, but these faults do not significantly offset mapped contacts. Bedding dips in volcanic units are typically 35°–65° to the southwest (Miller, 1970), suggesting that northeast-dipping normal faults, now concealed or obscure, are dominant.

Cross-section reconstruction of the Buckskin detachment fault changes the dip of the seismic reflectors beneath Cactus Plain and the layered metasedimentary tectonites of the Battleship Peak area so that both dip northeastward rather than southwestward (Fig. 13B). The basic geometry of the reconstruction is to restore the Battleship Peak area to 10–15 km beneath the hanging-wall block of the Buckskin-Rawhide detachment fault, while the base of tilted Miocene strata in the northwestern Bouse Hills retains its position at Earth’s surface. The reconstructed tectonites and seismic reflectors project up dip to the south toward the Quinn Pass shear zone and, farther south, to the thrust sheets of the southern Plomosa Mountains and deformed Paleozoic rocks of the Boyer Gap area (Fig. 2). This entire zone is readily interpreted as a north-dipping thrust zone that was part of the Maria fold and thrust belt (Reynolds et al., 1986; Spencer

Figure 13. Modern (A, C) and restored (B) cross-sections for the Plomosa Mountains area. See Figure 2 for location. “Pz” (blue) represents Paleozoic and less common Mesozoic metasedimentary rocks. (A) Features identified from mapping projected into the subsurface. Also shown are seismic reflectors from Figure 9 (red lines). (B) Restoration of extension and reconstruction of features shown in (A). Orocopia Schist is projected northeastward below Paleozoic tectonites now exposed at Battleship Peak. This requires a northeast-dipping upper contact for the schist in this area. Strata deposited in Miocene basins are illustrated above the land surface, which is intended to facilitate visualization of the reconstruction. In two small areas dotted lines and adjacent faults separate bedrock that is now eroded away. (C) Features shown in cross section (A) plus Orocopia Schist after Miocene tectonic extension. Note that Miocene intrusions are not shown in the subsurface but must be abundant.
The reconstructed cross section shows the metasedimentary tectonites along the Buckskin detachment-fault footwall from Battlebox Peak to Lincoln Ranch basin 20 km to the northeast (Fig. 13B; Bryant, 1995; Singleton et al., 2014). The marble, schist, and quartzite tectonites were first buried by thrust faulting to depths of amphibolite-facies metamorphism (Bryant and Wooden, 2008; Singleton et al., 2018). The tectonites are sheared at their base without a basal Cambrian quartzite unit resting depositionally on Proterozoic bedrock (Marshak and Vander Meulen, 1989; Singleton et al., 2018). As a result, the Mesozoic history of tectonic burial must be multi-staged with metasedimentary units forming thrust slivers before shearing along the Buckskin detachment fault (fig. 36 in Spencer and Reynolds, 1989b). During post-thrusting detachment faulting, the tectonites were displaced northeastward from the Battleship Peak area and smeared out along the detachment-fault footwall. This process, however, could have been an overprint of thrust-related deformation, with the distribution of tectonites primarily reflecting thrust-related deformation rather than Miocene detachment-related shearing and smearing.

Correlation of the Battleship Peak tectonites with the Quinn Pass shear zone and structural elements of the Maria fold and thrust belt farther south is consistent with the interpretation of Singleton et al. (2018) that the metasedimentary tectonites below the detachment fault northeast of Battlebox Peak are also an element of the Maria fold-thrust belt that localized the detachment fault during later extension. Furthermore, carbonate sheets and slivers within mylonitic quartz-feldspathic gneiss below the Buckskin-Rawhide detachment fault have been mapped in the Planet Peak area (Scott, 2004) and identified where mineralized at scattered locations across the Buckskin and Rawhide Mountains (Alamo, Squaw Peak, and Owen mineral districts of Spencer and Welty, 1989). The wide distribution of these small sheets and slivers, estimated by Singleton et al. (2018) to have extended over 25%–35% of the detachment-fault footwall, raises the possibility that the ~40-km-long lateral ramp on the north flank of the Buckskin and Rawhide Mountains (Fig. 1) is a consequence of detachment-fault localization in carbonates that were deeply buried by Mesozoic thrust faults. In this case, reduced fault friction in the strongest part of the quartzo-feldspathic upper crust would reduce minimum fault dip for stable sliding, resulting in footwall exhumation without tectonic extension of the overlying wedge (Singleton et al., 2018).

**Geodynamic Implications of Orocopia Schist**

On the restored cross section of Figure 13B, the upper boundary of the Orocopia Schist is a curved surface that dips northeastward beneath the Buckskin detachment-fault footwall and flattens southwestward beneath the southern Plomosa and Dome Rock Mountains. This geometry is indicated by the lack of exposures of the schist in these nearby areas. The inferred modern distribution of the schist is shown in Figure 13C, which was constructed by extending the schist as shown in Figure 13B to the modern extended geometry shown in Figure 13A. The cross section suggests that the schist and younger igneous intrusions now form all crust below ~5 km depth along the entire 50-km-long cross section. Alternatively, the schist could form diapirs or wedge-like intrusions or have some other complex geometric relationship with preexisting quartzo-feldspathic crust, although such complexities are not apparent in seismic reflection and refraction data that reveal generally subhorizontal layering (McCarthy et al., 1991). The inferred great lateral extent of the schist, consistent with underplating following tectonic ablation of the underside of the lithosphere (Spencer, 1996; Strickland et al., 2018), suggests that any crustal root created by tectonic shortening in the Maria fold-and-thrust belt was removed before its buoyancy could contribute to isostatic uplift of the Harcuvar core complex, and that core-complex exhumation and uplift had causes other than the buoyant root model of Coney and Harms (1984) and Spencer and Reynolds (1990b).

Although corrugation crests in the Harcuvar core complex are as high as 1700 m, by far most of the core complex and surrounding areas are at ~400–800 m elevation, with lower elevations (100–400 m) to the west in the Colorado River valley. Core-complex uplift and exhumation that yielded this generally low-relief topography could have resulted from deep, quartz- and mica-rich crust that behaved as a low viscosity fluid at tectonic rates and distance scales, combined with a detachment-fault footwall with low flexural strength. In other words, a flimsy detachment-fault footwall above a laterally extensive, tectonically fluid substrate would respond to tectonic denudation by rising to the approximate level of regional topography (Block and Royden, 1990; Wernicke, 1990, 1992). Uplift and exhumation would have occurred beneath an inclined detachment fault, with flattening of the fault footwall following a laterally migrating monoclinical flexure adjacent to, or beneath, the trailing end of the upper plate wedge (Wernicke and Axen, 1988; Buck, 1988; Spencer and Reynolds, 1991). Mobilization and uplift of Orocopia Schist beneath the core complex may have been aided by the early Miocene magmatic pulse and by high fluid pressures associated with magmatic heating and dehydration of the schist.

**CONCLUSIONS**

The early Miocene geologic history of the northern Plomosa Mountains and western Bouse Hills was characterized by a period of basin genesis and sedimentation, mafic volcanism, and rock-avalanche deposition, followed by a period of extensional faulting that broke up and tilted previously deposited strata and was associated with dacitic volcanism and minor sedimentation. The first period of basin genesis and subsidence occurred at ca. 20–23 Ma, as indicated by dates of interbedded tuffs. The second period occurred at ca. 15–19 Ma as indicated by fission-track dates of detachment-fault footwall rocks interpreted as reflecting tectonic denudation. Cross-section analysis indicates that an extension-parallel transect that was originally ~9.6 km wide is now ~21.7 km wide due to ~116% tectonic extension.

Critical-taper theory provides an explanation for delayed breakup of the extensional wedge above the Plomosa detachment fault. Early basin genesis and deposition of strata were followed by initiation of the Plomosa detachment fault and subsequent tilting of footwall and hanging-wall blocks. This tilting was associated with, or followed by, breakup of the hanging-wall wedge above the Plomosa detachment fault. Wedge breakup is attributed to tilting of the underlying detachment fault and the land surface so that the wedge geometry reached and occupied the extensional boundary of the critical-taper stable-sliding field.

Restoration of extension along a cross section through the northern Plomosa Mountains and adjacent areas to the northeast and southwest indicates that Paleozoic and Mesozoic strata that were deformed and metamorphosed by Mesozoic thrusting and thrust burial were aligned in a northeast-dipping zone before tectonic extension. This restoration aligns the Quinn Pass thrust zone with the Battleship Peak carbonate and calc-silicate tectonite in the southern Buckskin Mountains and the thrust faults and deformed Paleozoic and Mesozoic strata in the southern Plomosa Mountains and northern Dome Rock Mountains. The Quinn
Pass–Battlefield Peak metasedimentary tectonite zone projects downlap into a thin sheet of tectonites below the Buckskin detachment fault on the north flank of the Ives Peak arch. This geometry is consistent with the interpretation that the tectonite sheet was originally an element of a Mesozoic thrust zone, and that the Buckskin-Rawhide detachment fault was localized in carbonate tectonites where the tectonites formed weak zones in otherwise strong quartzo-feldspathic crust, as proposed by Singleton et al. (2018).

The presence of Oroopia Schist in the northern Plomosa Mountains strongly suggests that the lower and middle crust in the region was replaced by subduction complex during latest Cretaceous to Paleogene subcrustal tectonic erosion and schist underplating. This would have eliminated the buoyant crustal root that had been proposed as the cause of uplift of the Harcuvar core complex. We propose that a highly mobile deep crust combined with flexurally weak detachment-fault footwalls would be adequate to accommodate uplift of deep crustal rocks to approximately the level of the surrounding region, consistent with previous interpretations of core-complex genesis (Wernicke and Axen, 1988; Block and Ruyen, 1990; Wernicke, 1992). We suggest that the inferred high mobility of the deep crust was enhanced by the hydrous composition of subduction-complex rocks and by intrusion of abundant lower Miocene granitoids.

The buoyant crustal root hypothesis for core-complex exhumation was originally applied to the thickened axis of the Cordilleran orogen (Coney and Harms, 1984) without any obvious association of individual core complexes with local areas of exceptional thinning. This generally thickened crust, however, would be expected to have a highly mobile deep crust, with mobility proportional to the cube of the thickness of the channel in which deep crust flows (Kruse et al., 1991). It thus appears that core complexes are associated with the thickened axis of the Cordilleran orogen because of the high mobility of thick continental crust rather than because individual core complexes are associated with specific local areas of exceptional thinning. This may be true of other core complexes around the world.

ACKNOWLEDGMENTS
We thank Rob Scott (University of Tasmania) for an especially thorough review, and Lithosphere science editor Kurt Stüwe and an anonymous reviewer for useful suggestions regarding the focus of this paper. Field mapping by J. Spencer and D. Love in the northern Plomosa Mountains along the Plomosa-Copper peak traverse in 2013 was supported by the U.S. Geological Survey National Cooperative Geologic Mapping Program under STATEMAP grant G12AC20446. Mapping by E. Strickland and J. Singleton in the northern Plomosa Mountains was supported by U.S. Geological Survey 2016 EDMAP grant G16AC00142. The views and conclusions contained in this document are those of the authors and should not be interpreted as necessarily representing the official policies, either expressed or implied, of the U.S. government. This manuscript is submitted for publication with the understanding that the U.S. government is authorized to reproduce and distribute reprints for governmental use.

REFERENCES CITED
Howard, K.A., and John, B.E., 1984, Eocene crust extension along a rooted system of late stage 1:9,318.

Geological Society of America | LITHOSPHERE | Volume 10 | Number 6 | www.gsapubs.org


